Climatic and geomorphic controls on the erosion of terrestrial biomass from subtropical mountain forest

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[1] Erosion of particulate organic carbon (POC) occurs at very high rates in mountain river catchments, yet the proportion derived recently from atmospheric CO_2 in the terrestrial biosphere ($POC_{non\text{-}fossil}$) remains poorly constrained. Here we examine the transport of $POC_{non\text{-}fossil}$ in mountain rivers of Taiwan and its climatic and geomorphic controls. In 11 catchments we have combined previous geochemical quantification of POC source (accounting for fossil POC from bedrock), with measurements of water discharge (Q_w) and suspended sediment concentration over 2 years. In these catchments, $POC_{non\text{-}fossil}$ concentration (mg L^{-1}) was positively correlated with Q_w, with enhanced loads at high flow attributed to rainfall driven supply of $POC_{non\text{-}fossil}$ from forested hillslopes. This climatic control on $POC_{non\text{-}fossil}$ transport was moderated by catchment geomorphology: the gradient of a linear relation of $POC_{non\text{-}fossil}$ concentration and Q_w increased as the proportion of steep hillslopes (>35°) in the catchment increased. The data suggest enhanced supply of $POC_{non\text{-}fossil}$ by erosion processes which act most efficiently on the steepest sections of forest. Across Taiwan, $POC_{non\text{-}fossil}$ yield was correlated with suspended sediment yield, with a mean of 21 ± 10 tC km^{-2} yr^{-1}. At this rate, export of $POC_{non\text{-}fossil}$ imparts an upper bound on the time available for biospheric growth, of ~800 yr. Over longer time periods, $POC_{non\text{-}fossil}$ transferred with large amounts of clastic sediment can contribute to sequestration of atmospheric CO_2 if buried in marine sediments. Our results show that this carbon transfer should be enhanced in a wetter and stormier climate, and the rates moderated on geological timescales by the regional tectonic setting.

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1. Introduction

[2] The majority of organic carbon found at Earth's surface resides on the continents, with $\sim 2100 \times 10^{15}$ gC stored in soils and vegetation of the terrestrial biosphere and a further significant amount of fossil organic carbon contained within outcropping sedimentary rocks [*Sundquist*, 1993; *Sigman and Boyle*, 2000; *Holmén*, 2000]. Therefore, the physical erosion of the continents and the concomitant transfer of particulate organic carbon (POC) to the oceans by rivers is an important component of the global carbon cycle [*Ittekkot*, 1988; *Sarmiento and Sundquist*, 1992; *Meybeck*, 1993; *Ludwig et al.*, 1996; *Stallard*, 1998]. If this POC is derived from recently photosynthesized organic matter from the biosphere ($POC_{non\text{-}fossil}$), then its transfer represents the export of a fraction of terrestrial primary productivity [*Hilton et al.*, 2008a]. It can contribute to the geological sequestration of atmospheric CO_2 if $POC_{non\text{-}fossil}$ is buried in long-lived sedimentary deposits [*Berner*, 1982; *Hedges and Keil*, 1995; *Stallard*, 1998; *France-Lanord and Derry*, 1997; *Hayes et al.*, 1999]. The highest rates of POC transfer, which includes fossil POC from bedrock (POC_{fossil}), have been measured in small river catchments (<5,000 km^2) draining mountainous terrain [*Kao and Liu*, 1996; *Lyons et al.*, 2002; *Hilton et al.*, 2008b] where large amounts of clastic sediment are also mobilized and exported by mountain rivers [*Milliman and Syvitski*, 1992; *Hovius et al.*, 2000; *Dadson et al.*, 2003]. As a result, these catchments are thought to contribute disproportionately to the supply of POC to large fluvial systems [*Mayorga et al.*, 2005; *Bouchez et al.*, 2010; *Galy and Eglinton*, 2011], its export to the oceans [*Lyons et al.*, 2002], and effective carbon burial, promoted by

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rapid sediment accumulation in depocenters [*Canfield*, 1994; *Leithold and Hope*, 1999; *Burdige*, 2005; *Galy et al.*, 2007; *Brackley et al.*, 2010].

[3] Estimates of POC yields from mountain catchments often lump POC_{fossil} from bedrock with $POC_{non\text{-}fossil}$ eroded from the terrestrial biosphere. Despite the potential importance of erosion and burial of $POC_{non\text{-}fossil}$ from mountain catchments, quantitative constraints are lacking. Consequently it is difficult to evaluate the role of external factors (e.g., climate, tectonics) in this carbon transfer [cf. *West et al.*, 2005]. First, this is due to difficulties in the sampling of mountain rivers with flashy hydrographs [*Hicks et al.*, 2004a; *Dadson et al.*, 2005] over the full range of flow conditions under which POC is transported [e.g., *Blair et al.*, 2003; *Hilton et al.*, 2008b]. Second, the input of POC_{fossil} from exhumed sedimentary rocks remains unconstrained in many settings [*Lyons et al.*, 2002; *Gomez et al.*, 2003]. POC_{fossil} is intimately associated with clastic sediment [*Leithold et al.*, 2006] and its transfer in suspended load has been shown to be strongly linked to sediment yield [*Hilton et al.*, 2011a]. By limiting its oxidation, erosion of POC_{fossil} and its input to the fluvial system imparts its chemical composition to terrestrial sediments [*Blair et al.*, 2003; *Leithold et al.*, 2006; *Hilton et al.*, 2010] and its reburial has important implications for our understanding of the global carbon cycle [*Dickens et al.*, 2004; *Galy et al.*, 2008; *Hilton et al.*, 2011a]. However, it does not represent a transfer of recently sequestered atmospheric CO_2 and so must be distinguished from $POC_{non\text{-}fossil}$ in river sediments [*Kao and Liu*, 1996; *Galy et al.*, 2007; *Hilton et al.*, 2008a, 2008b, 2010].

[4] In order to examine the controls on $POC_{non\text{-}fossil}$ transport and quantify its rate of transfer, $POC_{non\text{-}fossil}$ concentration must be examined as a function of water discharge (Q_w, $m^3\ s^{-1}$). Only a handful of studies have achieved this, focusing on individual catchments to provide quantification of annual and flood-driven $POC_{non\text{-}fossil}$ transfer [*Kao and Liu*, 2000; *Hilton et al.*, 2008a; *Townsend-Small et al.*, 2008; *Hatten et al.*, 2012]. These studies have identified the importance of: i) runoff and runoff variability; ii) catchment geomorphic setting; iii) physical erosion rate; and iv) aboveground carbon stock for $POC_{non\text{-}fossil}$ transport and transfer. To understand better how these climatic, geomorphic and biological drivers operate, we require measurements of the fluvial transport of $POC_{non\text{-}fossil}$ (mg L^{-1}) and estimates of $POC_{non\text{-}fossil}$ yields (tC $km^{-2}\ yr^{-1}$) from multiple catchments across gradients in controlling variables.

[5] Here we focus on the role of climatic and geomorphic factors in the forested mountain belt of Taiwan, where organic carbon stocks are relatively uniform [*Chang et al.*, 2006; *West et al.*, 2011]. We have obtained hydrometric data (Q_w and suspended sediment concentration) and collected suspended sediment samples from 11 major rivers draining the Central Range mountains over two years. The abundance of POC_{fossil} in these samples has been quantified previously [*Hilton et al.*, 2010, 2011a] allowing, for the first time, an examination of the mobilization and transport of $POC_{non\text{-}fossil}$ from a subtropical mountain forest as a function of Q_w. Moreover, constraints on the prevalence of steep hillslopes in study catchments provides new insight into how $POC_{non\text{-}fossil}$ transfer is moderated by the erosion processes which supply $POC_{non\text{-}fossil}$ to the river channel. Finally, using suspended sediment yield, we assess the role of physical erosion rate on $POC_{non\text{-}fossil}$ export and examine its impact on the time available for development of the mountain biosphere and the implications for regional and global carbon cycles.

2. Study Area

2.1. Tectonic and Climatic Setting

[6] Taiwan is located at 22–25°N at the western edge of the Pacific Ocean. Mountain building is driven by collision between the Luzon Arc on the Philippine Sea Plate and the Eurasian continental margin since ~7 Ma [*Beyssac et al.*, 2007]. It has formed the Central Range, standing almost 4 km above sea level and 9 km above the nearby ocean floor. Bedrock rivers drain its steep topography to the Pacific Ocean and Taiwan Strait [*Dadson et al.*, 2003; *Kao and Milliman*, 2008] and have cut into metamorphosed Mesozoic and Cenozoic siliciclastic and carbonate rocks [*Ho*, 1986; *Hartshorn et al.*, 2002] which contain between 0.2 and 0.4 weight % of POC_{fossil} [*Kao and Liu*, 2000; *Hilton et al.*, 2010]. The climate is subtropical, with 2–4 m yr^{-1} of rainfall, most of which falls between June and October when tropical cyclones (typhoons) impact the island [*Wu and Kuo*, 1999; *Galewsky et al.*, 2006].

[7] The tectonic setting and climatic conditions combine to produce high physical erosion rates, on average of 3–7 mm yr^{-1} in the Central Range resulting in the export of 380×10^6 t yr^{-1} of suspended sediment to the ocean between 1970 and 1999 [*Dadson et al.*, 2003; *Fuller et al.*, 2003]. Much of this sediment derives from bedrock landslides that mobilize clastic sediment from steep hillslopes [*Hovius et al.*, 2000] and act to turnover forested hillslopes [*Hilton et al.*, 2008a, 2011b]. Physical erosion outpaces chemical weathering rate by a factor of 10^3 [*West et al.*, 2005; *Calmels et al.*, 2011] which limits soil development in Taiwan [*Tsai et al.*, 2001; *Ho et al.*, 2012]. Generally, typhoons trigger one or more large floods each year in river catchments and these hydrological events play a crucial role in sediment transfer [*Dadson et al.*, 2005; *Kao and Milliman*, 2008]. The high frequency of their occurrence provides an opportunity to monitor erosion and transfer of $POC_{non\text{-}fossil}$ over a large dynamic range of flow conditions while sampling over a relatively short (annual) period [*Kao and Liu*, 1996; *Hilton et al.*, 2008a].

2.2. Vegetation Cover and Catchment Characteristics

[8] The humid climate of Taiwan sustains vegetation throughout the Central Range, where forest reaches the highest ridge crests. The evergreen forest contains *Ficus*, *Machilus*, *Castanopsis*, *Quercus*, *Pinus*, *Tsuga*, and *Picea* [*Su*, 1984] and large areas of the mountain ecosystem are protected with logging monitored [*Lu et al.*, 2001]. The aboveground standing biomass of the mixed conifer-hardwood forest in Taiwan is $21.6 \pm 9.4 \times 10^3$ t km^{-2} [*West et al.*, 2011], representing an average organic carbon stock of $11 \pm 5 \times 10^3$ tC km^{-2}. Soils in Taiwan are relatively thin due to the rapid physical denudation rate [*Hovius et al.*, 2000; *Dadson et al.*, 2003], with the average base of the saprolite at ~0.8 m (n = 310) in a Central Range catchment [*Tsai et al.*, 2001; *Ho et al.*, 2012]. A-horizons are ~0.1 m thick and contain the majority of the organic matter [*Tsai et al.*, 2001], with surface soils (<0.1 m)

Table 1. Geomorphic Characteristics, Suspended Sediment Yield (SSY), and Mean Water Discharge (Q_w) for the Sampled Rivers Over the Study Period

River	Area (km^2)	Slope[a] (deg)	Area With Slope > 35°[a] (%)	SSY[b] (t km^{-2} yr^{-1})	σ SSY[b] (t km^{-2} yr^{-1})	Mean Q_w (m^3 s^{-1})
Linpien	310	30	23	2909	304	26
Hsiukuluan	1539	31	31	4061	1611	109
Laonung	812	34	43	4399	301	105
Wulu	639	31	37	10344	1445	51
LiWu	435	37	52	18571	4806	30
Heping	553	33	41	18704	5097	50
Chenyoulan	367	35	46	21064	1485	31
Choshui	2906	35	40	22798	1781	216
Hualien	1506	33	35	25292	10740	180
Yenping	476	31	36	58897	5422	70
Peinan	1584	31	31	72993	20302	125

[a]Median slope angle derived from 40 m DEM in ArcGIS.
[b]SSY and error on yield (σ SSY) from *Hilton et al.* [2011a].

beneath coniferous forest found to contain 7 ± 2 × 10^3 tC km^{-2} [*Chang et al.*, 2006]. The values of organic carbon stock are similar to averages of lowland tropical forests [*Dixon et al.*, 1994].

[9] The river catchments selected for study drain the Central Range and range in size from 310 km^2 to 2,906 km^2 (Table 1) covering ~30% of Taiwan's surface area. Upstream, the land use is dominated by mixed conifer-hardwood forest [*West et al.*, 2011; M. C. Hansen et al., Vegetation continuous fields MOD44B, 2001 Percent Tree Cover, Collection 4, 2006, http://glcf.umiacs.umd.edu/data/vcf/, hereinafter referred to as Hansen et al., online data set, 2006]. During the study period, the mean annual runoff was relatively constant between the catchments at 2.9 ± 0.2 m yr^{-1} (n = 11; ± standard error), suggesting no significant gradients in mean annual precipitation. However, within each catchment runoff variability was marked, with daily mean Q_w ranging over a factor 300 from ~0.1 to ~30 times the mean. In addition, the mean suspended sediment yield varied by up to a factor of 25 between catchments (Table 1), with a mean of 24,000 ± 7,000 t km^{-2} yr^{-1} (n = 11; ± standard error) [*Hilton et al.*, 2011a]. The study catchments also have variable geomorphic characteristics, which reflect the tectonic evolution of the mountain belt as well as the local bedrock geology [*Dadson et al.*, 2003; *Ramsey et al.*, 2007]. The distribution of hillslope angles in each catchment, a primary control on the rates of physical erosion processes in mountain topography [*Dietrich et al.*, 2003], varies notably. In most lithologies with pervasive jointing, hillslopes become disproportionately prone to failure at an angle of 30°–35°, with bedrock landslides most likely on the steepest sections of topography [*Burbank et al.*, 1996; *Clarke and Burbank*, 2010]. The rates of erosion by processes other than bedrock landslides (e.g., shallow landsliding, overland flow) also increase rapidly above this threshold [*Roering et al.*, 1999, 2001]. We therefore quantify the proportion of surface area with slopes >35° from a 40 m DEM of Taiwan [*Dadson et al.*, 2003] and find that this varies significantly among the studied catchments. In the Linpien catchment, located in the South West where relief is relatively low and Cenozoic inter-bedded sandstones and shales dominate the geology [*Ramsey et al.*, 2007; *Hilton et al.*, 2010], 23% of the catchment area has slopes >35° (Table 1). In the Liwu River in the North East, which is underlain by more competent, high-grade metamorphic rocks [*Ramsey et al.*, 2006; *Beyssac et al.*, 2007; *Hilton et al.*, 2010], very steep slopes are prevalent over 52% of the catchment area (Table 1).

3. Materials and Methods

3.1. Sample Collection, Processing and Geochemical Analyses

[10] Suspended sediment samples were collected at 11 gauging stations where water discharge (Q_w, m^3 s^{-1}) and suspended sediment concentration (SSC, mg L^{-1}) are routinely monitored. The details of our sampling methods have been described elsewhere [*Dadson et al.*, 2003; *Hilton et al.*, 2008a; *Kao and Milliman*, 2008; *Hilton et al.*, 2010]. In summary, rivers were sampled, on average, one to three times per month over two typhoon seasons in 2005 and 2006 (Figure 1a and Table 1). The Liwu River was sampled during 2004 in a similar manner, but suspended load was also collected at a higher, daily frequency during specific typhoon floods [*Hilton et al.*, 2008a]. Given the turbulence of these rivers at the sampling site, our samples are representative of the suspended sediment carried by the rivers [*Lupker et al.*, 2011]. The maximum grain size of $POC_{non\text{-}fossil}$ in these samples was found to be ~500 μm [*Hilton et al.*, 2010]. The transfer of coarse woody debris (CWD), while potentially important [*West et al.*, 2011], was not quantified in this study.

[11] The concentration of suspended $POC_{non\text{-}fossil}$ (mg L^{-1}) was determined following inorganic carbon removal and analysis of the organic carbon concentration of the suspended load (C_{org}, %), the nitrogen to organic carbon ratio (N/C) and the stable isotopes of organic carbon ($\delta^{13}C_{org}$, ‰) by a Costech Elemental Analyzer coupled via Conflo-III to a MAT-235 stable isotope mass spectrometer. The fraction of non-fossil POC (F_{nf}) was quantified using N/C and $\delta^{13}C_{org}$ and an end-member mixing analysis for each sample, detailed by *Hilton et al.* [2010]. $POC_{non\text{-}fossil}$ concentration (mg L^{-1}) for each sample was determined as the product of SSC, C_{org} and F_{nf}. *Hilton et al.* [2010] found F_{nf} to be a reliable predictor to correct for fossil POC input when tested against independent constraint from measurements of ^{14}C content in 9 samples from the Liwu River. This is an appropriate test catchment for the mixing model as it comprises geological formations spanning the full range in POC_{fossil} compositions found in the mountain belt [*Hilton et al.*, 2010]. F_{nf} was found to have an average precision of 0.09 and

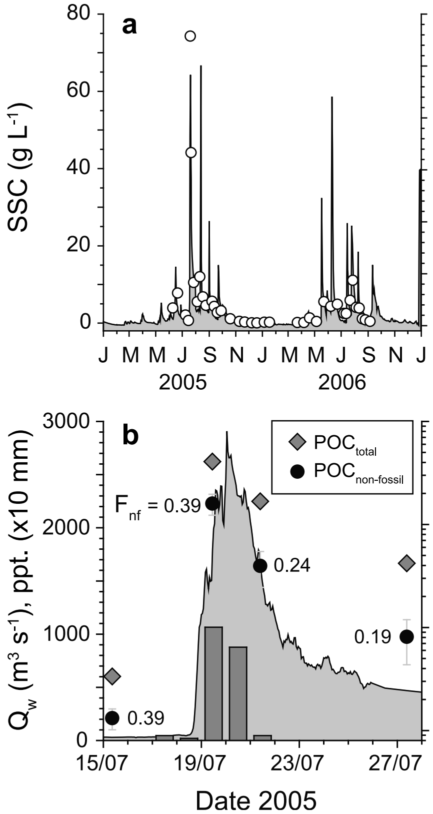

Figure 1. Hydrometric data and samples collected by the Water Resources Agency, Taiwan, for this study. (a) Daily average water discharge (Q_w, m^3 s^{-1}, filled gray curve) and measured suspended sediment concentration (SSC, mg L^{-1}) of samples (circles) from the Peinan River in 2005 and 2006. (b) Detail showing hourly water discharge (Q_w, filled gray curve) for the Peinan River and daily precipitation totals (ppt. $\times$ 10 mm, dark gray bars) for Taitung at the gauging station during Typhoon Haitang. Total POC concentration (POC_{total}, mg L^{-1}, gray diamonds) which includes fossil POC, fraction non-fossil (F_{nf}) and POC derived from vegetation and soil ($POC_{non\text{-}fossil}$, mg L^{-1}, black circles) are shown.

represents the largest source of uncertainty in our analysis of $POC_{non\text{-}fossil}$ transfer. The error on $POC_{non\text{-}fossil}$ concentration was highest in samples where $F_{nf} < 0.10$, with an average error of 50% across the catchments (n = 11). Weighted by Q_w, errors on $POC_{non\text{-}fossil}$ were lower on average (44%, n = 11) because F_{nf} was typically >0.2 at high flow (Figure 2) and the maximum absolute uncertainty in $POC_{non\text{-}fossil}$ concentration was 40 mg L^{-1}, ~25% of the calculated concentration in that sample (160 mg L^{-1}). Despite these limitations, F_{nf} provides robust constraint on the erosion of soil and vegetation POC, with the reported errors much smaller than the measured range in $POC_{non\text{-}fossil}$ concentration over three orders of magnitude (Figure 3).

3.2. $POC_{non\text{-}fossil}$, Q_w and Quantification of Yields

[12] Relationships between $POC_{non\text{-}fossil}$ concentration (mg L^{-1}) and Q_w in small mountain rivers have previously been described by power law [*Hilton et al.*, 2008a; *Hatten et al.*, 2012] and linear [*Townsend-Small et al.*, 2008] relationships. These relationships can be used to compare the transport of $POC_{non\text{-}fossil}$ in different catchments during similar hydrological conditions, using the mean Q_w (Q_{mean}) to normalize Q_w (Q_w/Q_{mean}). To date, power laws relations of $POC_{non\text{-}fossil}$ and Q_w have been fitted either to data in catchments where $POC_{non\text{-}fossil}$ dominates the total POC load [e.g., *Hatten et al.*, 2012] or where F_{nf} has been quantified by ^{14}C measurements [e.g., *Hilton et al.*, 2008a], i.e., when the error on each $POC_{non\text{-}fossil}$ measurement was negligible. This does not apply in our case due to uncertainty on F_{nf} [*Hilton et al.*, 2010]. In the majority of our 11 catchments, least squares best fits of power laws were not statistically significant, which may partly reflect the reported errors on $POC_{non\text{-}fossil}$ in this study. Instead, a linear relationship was quantified with slope (m-$POC_{non\text{-}fossil}$) and intercept (c-$POC_{non\text{-}fossil}$). Statistical analyses were carried out in Origin Pro™.

[13] Power law rating curves can be used to quantify the yield of particulate constituents. Suspended sediment yield (SSY, t km^{-2} yr^{-1}) was quantified by *Hilton et al.* [2011a] using rating curves between Q_w and SSC, then applied to the daily record of Q_w, for each catchment between 2005 and 2007 (2004 for the Liwu River) (Table 1). SSY was also quantified using water discharge-weighted mean SSC as described elsewhere [*Walling and Webb*, 1981; *Ferguson*, 1987]. This flux-weighted method (SSY_{fw}) can provide robust quantification of river loads in the absence of a power law rating curve [*Ferguson*, 1987]. It was, therefore, applied to estimate $POC_{non\text{-}fossil}$ yields (tC km^{-2} yr^{-1}) in each catchment.

4. Results

4.1. Fluvial Transport of $POC_{non\text{-}fossil}$

[14] Suspended sediments were collected over a large range in Q_w, with Q_w/Q_{mean} at the time of sampling ranging from ~0.1 to ~30 in catchments (Figure 2). Over this range, F_{nf} varied between 0 and ~0.8, the highest values occurring during low flows with $Q_w/Q_{mean} < 3$. For these flows, there

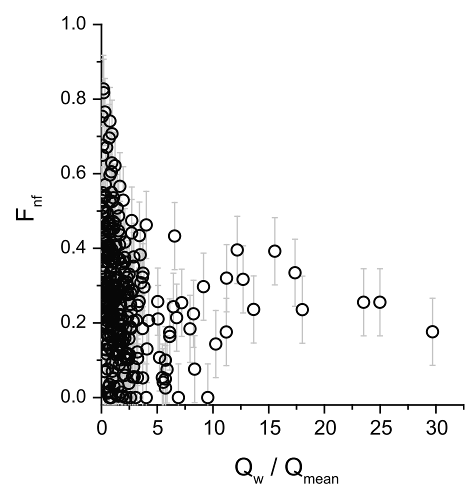

Figure 2. Fraction non-fossil POC (F_{nf}) versus water discharge (Q_w) normalized to the mean inter-annual water discharge (Q_{mean}) for all samples across the study catchments. Whiskers show errors on F_{nf}.

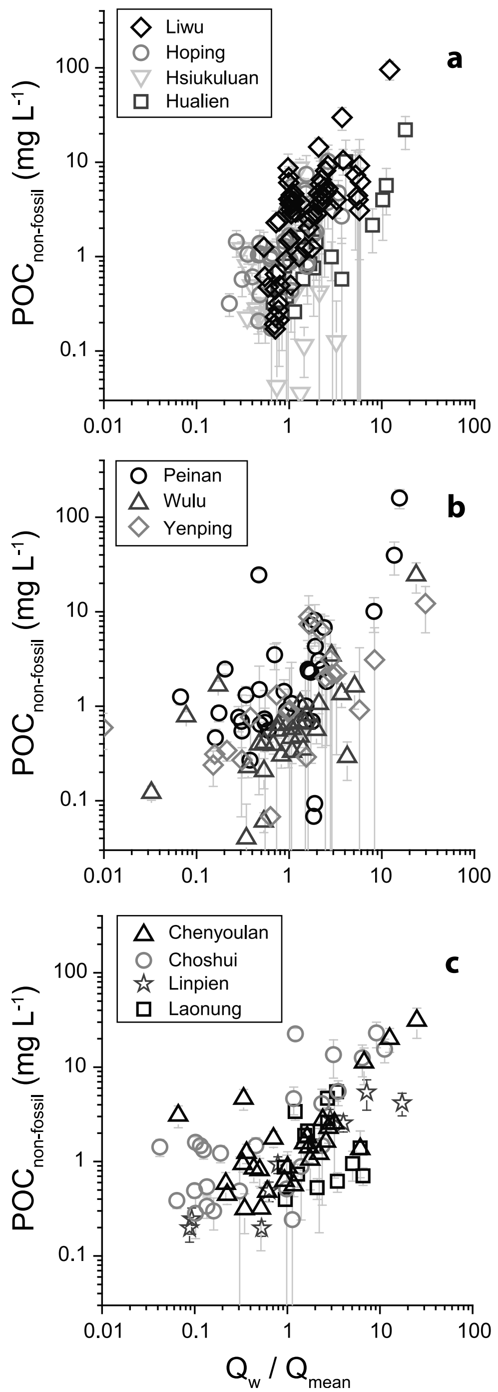

Figure 3. Relationship between normalized water discharge (Q_w/Q_{mean}) and $POC_{non\text{-}fossil}$ concentration (mg L^{-1}) in mountain rivers draining: (a) the North East, (b) the South East, and (c) the West of the Central Range Taiwan. Whiskers show errors on $POC_{non\text{-}fossil}$ concentration.

was a negative correlation between F_{nf} and Q_w/Q_{mean} ($r = -0.31$; $P < 0.0001$; n = 273) and a negative correlation between C_{org} and Q_w/Q_{mean} ($r = -0.40$; $P < 0.0001$; n = 273). However, for larger events during floods with $Q_w/Q_{mean} > 3$, there was no evidence for a decrease in F_{nf} with increased Q_w ($r = 0.2$; $P = 0.2$; n = 52; Figure 2), nor of any dilution of C_{org} ($r = 0.02$; $P = 0.8$; n = 52) across the sample set.

[15] Measured $POC_{non\text{-}fossil}$ concentrations (mg L^{-1}) ranged over three orders of magnitude (Figure 3) to a maximum of 160 ± 40 mg L^{-1}. This covers the range of previously reported concentrations in Taiwanese rivers and small mountain rivers elsewhere [*Hilton et al.*, 2008a; *Hatten et al.*, 2012]. The lack of a decrease in F_{nf} and C_{org} at high Q_w (Figure 2) resulted in a lack of dilution of $POC_{non\text{-}fossil}$ concentration (mg L^{-1}) (Figure 3) as suspended load mass increased with discharge [*Hilton et al.*, 2011a]. A strong positive correlation between $POC_{non\text{-}fossil}$ concentration and Q_w/Q_{mean} exists ($r = 0.49$; $P < 0.0001$; n = 325) which contrasts previous results from non-mountainous catchments [cf. *Ludwig et al.*, 1996; *Stallard*, 1998]. The positive correlation held in all but two of the sampled catchments (Figure 3 and Table 2), its gradient (m-$POC_{non\text{-}fossil}$) varying from 0.27 ± 0.08 in the Linpien River to 6.43 ± 0.78 in the Peinan River. The intercept (c-$POC_{non\text{-}fossil}$) varied between −5.0 ± 3.0 mg L^{-1} in the Peinan River to 1.1 ± 1.2 mg L^{-1} in the Choshui River.

[16] The sampling strategy did not specifically target floods caused by tropical cyclones [cf. *Goldsmith et al.*, 2008; *Hilton et al.*, 2008a] because of the logistical difficulties and hazards associated with these events. However, four samples were collected during Typhoon Haitang (onset 19 July 2005, flood peak at 01:00 20 July 2005) in the Peinan River. At the peak of the flood there was enhanced $POC_{non\text{-}fossil}$ transport at high Q_w (Figure 1b), confirming the observations made previously in other Taiwanese rivers [*Hilton et al.*, 2008a] and in flood deposits of the Waipaoa River, New Zealand [*Gomez et al.*, 2010].

4.2. Particulate Yields

[17] Across the area covered by the 11 catchments, the average $POC_{non\text{-}fossil}$ yield, estimated using the discharge-weighted mean $POC_{non\text{-}fossil}$ concentration, was 21 ± 10 tC km^{-2} yr^{-1}. Over the study period $POC_{non\text{-}fossil}$ yields varied from 1.2 ± 1.0 tC km^{-2} yr^{-1} in the Hsiukuluan River in central east Taiwan, to 74 ± 22 tC km^{-2} yr^{-1} in the Peinan River to the south (Table 2). The Peinan River yield is among the highest ever recorded for a multiannual average. $POC_{non\text{-}fossil}$ was approximately 30% of the total POC load exported by these mountain rivers, with POC_{fossil} making up the remaining part [*Hilton et al.*, 2010] and contributing, on average, 82 tC km^{-2} yr^{-1} [*Hilton et al.*, 2011a].

[18] $POC_{non\text{-}fossil}$ yields were strongly correlated with SSY over three orders of magnitude (Figure 4a), which was not the consequence of varying drainage area. Using this, we can compare the published SSY over the study period from power law rating curves (Table 1) [*Hilton et al.*, 2011a] with those derived from the same flux-weighted method, SSY_{fw}, to determine whether the discharge-weighted estimation of $POC_{non\text{-}fossil}$ yield is a robust method. The two SSY estimates are strongly, linearly correlated by $SSY_{fw} = 0.74 \pm 0.04*SSY$ ($r^2 = 0.96$; $P < 0.0001$; n = 11), suggesting that the $POC_{non\text{-}fossil}$ yields estimated by flux-weighting [*Ferguson*, 1987] are robust. However, the SSY_{fw} are on average 21% lower than published, rating curve-derived SSY (Table 2). This is because the flux-weighted method does not fully account for the role of very large floods in the annual hydrograph [*Ferguson*, 1987], for example during tropical-cyclones in Taiwan [*Dadson et al.*, 2005]. This is confirmed by the observation that the discharge-weighted

Table 2. $POC_{non\text{-}fossil}$ Transport and Transfer in the Study Catchments[a]

River	m-$POC_{non\text{-}fossil}$	σ m-$POC_{non\text{-}fossil}$	c-$POC_{non\text{-}fossil}$ (mg L^{-1})	σ c-$POC_{non\text{-}fossil}$ (mg L^{-1})	SSY_{fw} (t km^{-2} yr^{-1})[b]	Average F_{nf}[c]	$POC_{non\text{-}fossil}$ yield (tC km^{-2} yr^{-1})[d]	σ $POC_{non\text{-}fossil}$ yield (tC km^{-2} yr^{-1})
Linpien	0.27	0.08	0.90	0.54	1546	0.32	2.8	0.8
Hsk	nd	nd	nd	nd	2837	0.25	1.2	1.0
Laonung	nd	nd	nd	nd	3161	0.41	4.3	1.1
Wulu	1.00	0.05	−0.58	0.22	18603	0.26	13.8	4.8
LiWu	4.71	0.53	−4.16	1.45	8460	0.33	6.8	2.7
Heping	1.49	0.36	0.32	0.52	10434	0.23	9.3	4.4
Chenyoulan	1.27	0.08	−0.05	0.43	18898	0.26	19.6	6.8
Choshui	1.66	0.34	1.05	1.16	16800	0.30	20.8	7.1
Hualien	0.87	0.13	−0.65	0.66	19420	0.22	13.8	7.8
Yenping	0.37	0.08	1.27	0.56	48702	0.16	23.4	18.4
Peinan	6.44	0.78	−4.96	2.97	49882	0.36	74.4	22.3

[a]Here nd indicates linear fit between Q_w/Q_{mean} and $POC_{non\text{-}fossil}$ concentration was not statistically significant and so parameters were not determined.
[b]Flux-weighted SSY for study period.
[c]Flux-weighted average F_{nf}.
[d]Flux-weighted $POC_{non\text{-}fossil}$ yield and error on yield (σ).

$POC_{non\text{-}fossil}$ yield for the Liwu River (for 2004) was 6.8 ± 2.7 tC km^{-2} yr^{-1}, which is lower than previous estimate of $POC_{non\text{-}fossil}$ yield during Typhoon Mindulle in 2004 [*Hilton et al.*, 2008a] of 13 tC km^{-2} derived with a rating curve (Figure 4a). Aiming to examine the variability in $POC_{non\text{-}fossil}$ yield between catchments (as a function of geomorphic characteristics and physical erosion rate), we have not applied any correction for these underestimations of $POC_{non\text{-}fossil}$. Instead, we suggest that the $POC_{non\text{-}fossil}$ yields reported here are internally consistent, but are likely to be conservative.

[19] Over the study period, the combined export from the monitored catchments was 0.21 ± 0.04 × 10^6 tC yr^{-1} of $POC_{non\text{-}fossil}$ (Figure 4b). Assuming a yield of 21 ± 10 tC km^{-2} yr^{-1} across Taiwan's mountain forest (22,665 km^2), the corresponding $POC_{non\text{-}fossil}$ flux from the Taiwan orogen to the ocean in suspended sediment was 0.5 ± 0.2 × 10^6 tC yr^{-1}. To determine whether the measured yields are representative of a longer-term (decadal) export, we note that SSY over the sampling period (mean 24,000 ± 7,000 t km^{-2} yr^{-1}, ± standard error) were similar to those estimated in the same catchments by *Dadson et al.* [2003] over three decades, 1970–1999 (mean 22,000 ± 4,000 t km^{-2} yr^{-1}, ± standard error). In view of the strong correlation of SSY and $POC_{non\text{-}fossil}$ yields (Figure 4a) this suggests that the $POC_{non\text{-}fossil}$ yields are likely to be a representative, albeit conservative for reasons previously stated, estimate of the longer term $POC_{non\text{-}fossil}$ transfer.

5. Discussion

5.1. Fluvial Transport of $POC_{non\text{-}fossil}$: Capacity and Supply

[20] Our results demonstrate that C_{org} and F_{nf} do not decrease at high Q_w (Figure 2) and thus that $POC_{non\text{-}fossil}$ is not diluted at the peak of large flood events (Figure 1b). This leads to a positive correlation between $POC_{non\text{-}fossil}$ concentration and Q_w/Q_{mean} (Figure 3) which is analogous to that commonly observed between Q_w/Q_{mean} and SSC in mountain rivers [*Hovius et al.*, 2000; *Fuller et al.*, 2003; *Hicks et al.*, 2004a; *Kao and Milliman*, 2008; *Hovius et al.*, 2011]. For clastic sediment, SSC increase with Q_w is often attributed to variability in: i) the capacity of the river to transport sediment as suspended load; and ii) the supply of suspendable sediment (sand, silt and clay) to the river channel. In mountain rivers a third factor may also be important, namely the production of suspended sediment by pebble abrasion at high levels of bed shear stress and associated bed load transport [*Attal and Lavé*, 2009]. We hypothesize that these factors also control $POC_{non\text{-}fossil}$ transport and examine their potential roles herein.

[21] The capacity of a river to entrain and transport fine sediment increases with water flow velocity and turbulence [*Garcia and Parker*, 1991]. Given the restricted channel geometry in bedrock rivers [*Turowski et al.*, 2008], capacity is likely to increase with Q_w. Turbulent mixing, typical of mountain river channels with large scale bed roughness, may also increase the entrainment rate and transport capacity of the flow [*Jackson*, 1976]. $POC_{non\text{-}fossil}$ should be less dense than the accompanying mineral sediment load, even when waterlogged [*Buxton*, 2010], causing its propensity for entrainment and transport to increase rapidly with Q_w [*Hamm et al.*, 2011]. However, in five of the catchments we observe negative values for c-$POC_{non\text{-}fossil}$, the linear intercept between $POC_{non\text{-}fossil}$ concentration and Q_w (Table 2). The physical meaning of a negative intercept implies either a threshold for motion for $POC_{non\text{-}fossil}$, which may be the case for coarse woody debris (CWD) [*West et al.*, 2011; *Wohl*, 2011] but seems unlikely for fine $POC_{non\text{-}fossil}$ [*Hamm et al.*, 2011], or a limit on the transport of $POC_{non\text{-}fossil}$ in river channels imposed by its supply. River channels in Taiwan are characterized by a lack of vegetation due to frequent flooding preventing colonization by plants [*Hartshorn et al.*, 2002] and therefore the supply of $POC_{non\text{-}fossil}$ must originate from forested hillslopes.

[22] The rate at which geomorphic processes erode the landscape are known to depend on the steepness of the topography on which they act [*Roering et al.*, 2001], and high rates of physical erosion by landsliding and overland flow are therefore expected to occur in Taiwan. Overland flow preferentially mobilizes loose material and $POC_{non\text{-}fossil}$ from surface soils [*Gomi et al.*, 2008]. Bedrock landslides can remove entire tracts of mountain forest and soil, harvesting the whole biomass and mixing it with POC_{fossil} [*Hilton et al.*, 2008b; *West et al.*, 2011; *Hilton et al.*, 2011b]. The influence of supply on $POC_{non\text{-}fossil}$ transport can be examined using

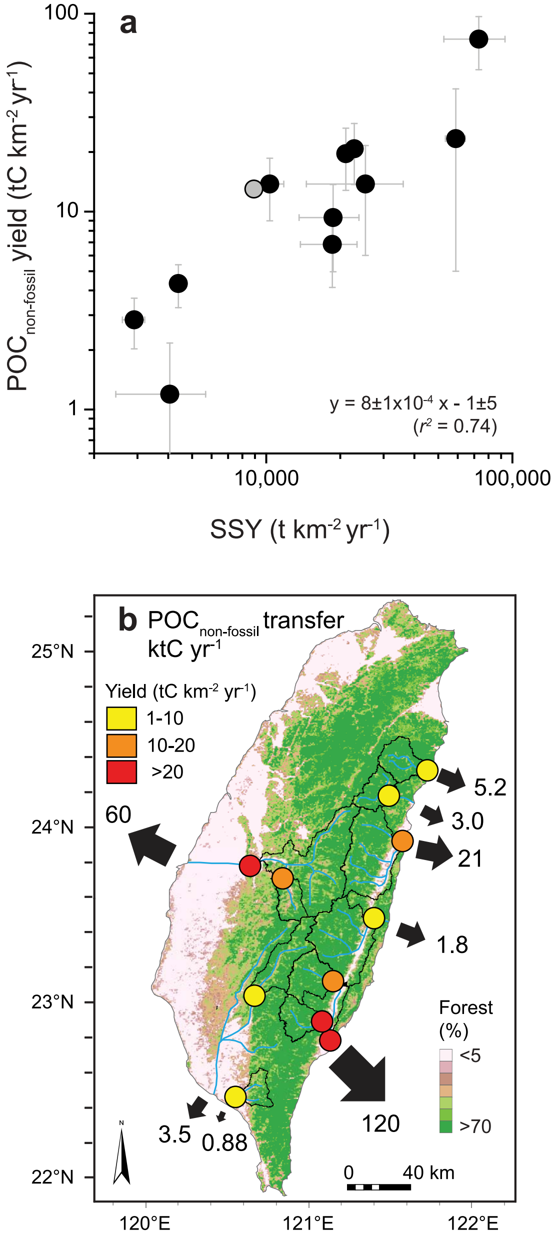

Figure 4. (a) Suspended sediment yield versus the $POC_{non\text{-}fossil}$ yield for the 11 Taiwanese catchments during the study period. Grey circle shows the published yields for Typhoon Mindulle in the Liwu River [*Hilton et al.*, 2008a] and whiskers show propagated errors. (b) $POC_{non\text{-}fossil}$ transfer (ktC yr^{-1}) to the ocean from Taiwan from the sampled catchments over the study period. $POC_{non\text{-}fossil}$ yields (tC km^{-2} yr^{-1}) shown (shaded circle) for each catchment. Forest cover (%) is shown derived from the Vegetation Continuous Fields product (Hansen et al., online data set, 2006).

the hysteresis of $POC_{non\text{-}fossil}$ and Q_w during individual flood events, as documented by *Hilton et al.* [2008a]. That study demonstrated that after several hours of sustained rainfall, enhanced $POC_{non\text{-}fossil}$ concentrations were observed across a range in Q_w when compared to dry intervals. Rainfall activates geomorphic processes of overland flow and landsliding and leads to efficient hillslope-channel coupling and the supply of $POC_{non\text{-}fossil}$. In addition, at high flood stage the river has capacity to transport CWD [*West et al.*, 2011] the mechanical attrition of which may also enhance $POC_{non\text{-}fossil}$ concentrations in the river suspended load [cf. *Attal and Lavé*, 2009]. In contrast, during periods without substantial rainfall, supply from hillslopes is minimal and $POC_{non\text{-}fossil}$ is likely to be sourced from channels, where bed sediments are typically dominated by POC_{fossil} [*Hilton et al.*, 2010]. Thus, $POC_{non\text{-}fossil}$ concentrations are lower for similar hydraulic conditions [*Hilton et al.*, 2008a].

[23] Organic carbon measurements on samples collected during the flood caused by Typhoon Haitang in the Peinan River are consistent with these observations [*Eglinton*, 2008]. Measured precipitation on 19 July 2005 totaled 110 mm in Taitung (Figure 1b) near to the gauging station (22.76°N, 121.15°E, data from the Central Weather Bureau, Taiwan, http://www.cwb.gov.tw/). On that day, the sample collected 14 h prior the peak of the flood, on the steep rising limb, had a $POC_{non\text{-}fossil}$ concentration of 160 ± 40 mg L^{-1} with F_{nf} = 0.39 ± 0.09. 32 h after the flood peak (09:40 21 July 2005), $POC_{non\text{-}fossil}$ concentration had dropped by 75% to 40 ± 15 mg L^{-1} (F_{nf} = 0.24 ± 0.09) despite only a slight decrease (~10%) in Q_w/Q_{mean} from 16 to 14. The marked drop in $POC_{non\text{-}fossil}$ concentration was co-incident with the cessation of heavy precipitation over the catchment (Figure 1b). These results demonstrate that while landsliding and overland flow are moderated by slope angle [*Dietrich et al.*, 2003], their temporal occurrence is stochastic [*Benda and Dunne*, 1997; *Hovius et al.*, 2000]. As a result, the fluvial transport of fine $POC_{non\text{-}fossil}$ may vary at a given transport capacity (Q_w) due to the specific timing and location of $POC_{non\text{-}fossil}$ supply to the river. This explanation is also consistent with the observed variability in $POC_{non\text{-}fossil}$ concentration for individual catchments (Figure 3) and confirms the importance of $POC_{non\text{-}fossil}$ supply during rainfall [*Hilton et al.*, 2008a], when erosion processes efficiently couple forested hillslopes to the river channel.

[24] The relative importance of the $POC_{non\text{-}fossil}$ supply processes identified here (overland flow, bedrock landslides, mechanical attrition) remains an avenue for future research. However, the observed lack of F_{nf} decrease with increasing Q_w provides some insight (Figure 2). As established, bedrock landslides are ubiquitous in Taiwan [e.g., *Lin et al.*, 2008] and known to be crucial for delivering clastic sediment to river networks at the peak of floods [*Hovius et al.*, 2000; *Fuller et al.*, 2003; *Dadson et al.*, 2005; *Hilton et al.*, 2008a]. However, erosion of POC by this process can decrease F_{nf} (decrease $POC_{non\text{-}fossil}$:POC_{fossil} ratio) at times of high sediment delivery. As the surface area of a bedrock landslide increases (i.e., its $POC_{non\text{-}fossil}$ erosion) it is known that its volume (i.e., sediment and POC_{fossil} erosion) increases as a power law with an exponent >1.2 [*Guzzetti et al.*, 2009; *Larsen et al.*, 2010], implying large landslides can dig deeper and reduce F_{nf} [*Hilton et al.*, 2008b]. Therefore, the observation of elevated F_{nf} during high flow (Figures 1b and 2) implies supply of $POC_{non\text{-}fossil}$ by a process other than deep bedrock landslides. Mobilization of surface materials by overland flow, and mechanical attrition of CWD do not contribute POC_{fossil}. One or both of these processes must contribute significantly to $POC_{non\text{-}fossil}$ fluxes in floods. These considerations support conclusions from the Western Southern Alps, New Zealand. There,

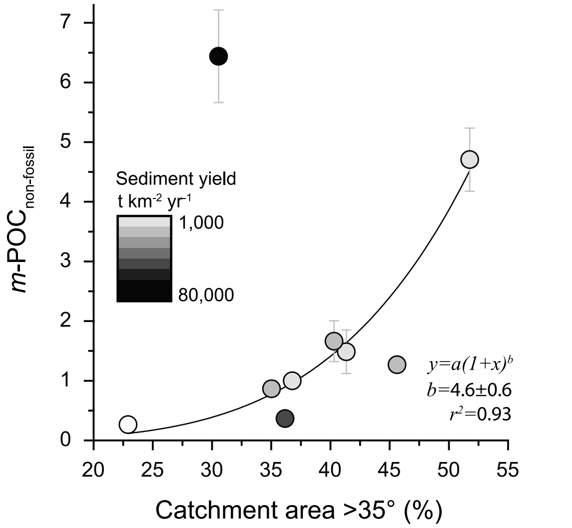

Figure 5. The gradient of the linear relationship between POC$_{non\text{-}fossil}$ and Q_w/Q_{mean} (Figure 3) for catchments which returned a significant fit (m-POC$_{non\text{-}fossil}$, Table 2) plotted against the proportion of catchment area with slope angles >35°. Shading of each point reflects the suspended sediment yield (Table 1). A nonlinear fit is shown to 8 of the catchments excluding the Peinan River.

decadal estimates of landslide-driven POC$_{non\text{-}fossil}$ yield were lower than estimates of fluvial export, requiring additional processes of POC$_{non\text{-}fossil}$ supply from the mountain hillslopes [*Hilton et al.*, 2011b].

5.2. Enhancement of POC$_{non\text{-}fossil}$ Transport

[25] Rainfall-driven changes in erosional supply underlie a strong climatic control on the mobilization and transport of POC$_{non\text{-}fossil}$ (Figure 3), which should have a similar expression in each catchment. However, it is clear that the positive relationship between POC$_{non\text{-}fossil}$ concentration and Q_w/Q_{mean} is not constant for Taiwanese Rivers. This is articulated in the range in gradients of the linear best fit to the data (m-POC$_{non\text{-}fossil}$), from 0.27 ± 0.08 to 6.43 ± 0.78 (Table 2). m-POC$_{non\text{-}fossil}$ can be viewed as an enhancement factor, with a steeper gradient reflecting increased loading of POC$_{non\text{-}fossil}$ across a range of hydrological conditions. As established previously (Section 5.1), supply is likely to be the main control on the variability in POC$_{non\text{-}fossil}$ concentration, rather than transport capacity in these rivers. Thus, enhancement should relate primarily to the efficiency of erosion processes delivering POC$_{non\text{-}fossil}$ from hillslopes to channels.

[26] The Taiwanese rivers have a positive trend between m-POC$_{non\text{-}fossil}$ and the area of the catchment with steep slopes above typical thresholds for mass wasting and erosion processes (>35°) (Figure 5). Between the Linpien River (Figure 3c) and the Liwu River (Figure 3a) the trend is nonlinear (n = 8). Such a trend is consistent with the mechanics of the geomorphic processes responsible for POC$_{non\text{-}fossil}$ supply [*Gomi et al.*, 2008; *West et al.*, 2011; *Hilton et al.*, 2011b]. Landsliding and overland flow processes are both stochastic and their rates of occurrence are a nonlinear, threshold functions of slope and runoff [*Benda and Dunne*, 1997; *Roering et al.*, 1999; *Hovius et al.*, 2000; *Dietrich et al.*, 2003]. Steepening the topography of a catchment should increase the rate of POC$_{non\text{-}fossil}$ supply, but only once hydrological thresholds are surpassed. This explains both the increase in POC$_{non\text{-}fossil}$ with Q_w (Figure 3) and enhanced rate of POC$_{non\text{-}fossil}$ supply when steep slopes contribute more importantly to the catchment hypsometry (Figure 5).

[27] The Peinan River, in the southwest of Taiwan, has an m-POC$_{non\text{-}fossil}$ of 6.43 ± 0.78 and lies significantly off the trend in the data set (Figure 5). To explain the higher loads of POC$_{non\text{-}fossil}$ in this catchment, we note that it also has had a very high suspended sediment yield for the study period, over the last four decades [*Dadson et al.*, 2003] and when compared to its mountain headwaters in the Wulu and Yenping catchments (Table 1 and Figure 4b). This may relate to active tectonic deformation of Pleistocene-Recent sediments in the Longitudinal Valley [*Ho*, 1986]. While the Wulu and Yenping mountain tributaries are located upstream (Figure 4b), the Peinan trunk river has cut into these recently uplifted, poorly consolidated sediments which contain POC$_{non\text{-}fossil}$ [*Shyu et al.*, 2006; *Ramsey et al.*, 2007]. Supply of clastic sediment and POC$_{non\text{-}fossil}$ from these deposits provides a mechanism to enhance fluvial POC$_{non\text{-}fossil}$ concentration across all Q_w (Figure 3b) and increase both the SSY and POC$_{non\text{-}fossil}$ yield. Cannibalism of young, uplifted foreland deposits may be an important mechanism by which POC$_{non\text{-}fossil}$ is re-mobilized in larger fluvial systems exiting active mountain belts [*Bouchez et al.*, 2010; *Galy and Eglinton*, 2011].

5.3. Export of POC$_{non\text{-}fossil}$ From Subtropical Mountain Forest

[28] The climatic (Figures 1b and 3) and geomorphic factors (Figure 5) that influence transport of POC$_{non\text{-}fossil}$ in Taiwan's mountain rivers also affect their clastic load [*Dietrich et al.*, 2003; *Dadson et al.*, 2003; *Hicks et al.*, 2004a; *Galewsky et al.*, 2006; *Kao and Milliman*, 2008]. As a result, a strong positive relationship exists between POC$_{non\text{-}fossil}$ yield and suspended sediment yield over two orders of magnitude in this mountain belt (Figure 4a). The data show no evidence for dilution of POC$_{non\text{-}fossil}$ yields at very high physical erosion rates. The average rate of POC$_{non\text{-}fossil}$ transfer of 21 ± 10 tC km^{-2} yr^{-1} represents an export of 0.12 ± 0.08% yr^{-1} of the total organic carbon stock in vegetation and soil, of 11 ± 5 × 10^3 tC km^{-2} and 7 ± 2 × 10^3 tC km^{-2}, respectively [*Chang et al.*, 2006; *West et al.*, 2011]. These export rates are high when compared to rates of geomorphic disturbance in mountain forest. In the western Southern Alps, New Zealand, bedrock landslides disturb forested surfaces at a rate 0.03% yr^{-1} [*Hilton et al.*, 2011b] and in Central America, disturbance rates are 10 times lower [*Restrepo and Alvarez*, 2006]. However, the POC$_{non\text{-}fossil}$ export rates here are likely to include important input from non-bedrock landslide inputs (overland flow, mechanical attrition of CWD) as previously discussed.

[29] The fluvial POC$_{non\text{-}fossil}$ export from the mountain forest has important implications for carbon cycling at the regional scale. In the absence of other output fluxes (e.g., respiration), it sets a bound on the amount of time available for organic matter to age in the landscape ($\tau_{non\text{-}fossil}$, yr). At a depletion-rate of 0.12 ± 0.08% yr^{-1}, physical erosion sets a timescale of 830 ± 530 yr for the aging of the organic carbon

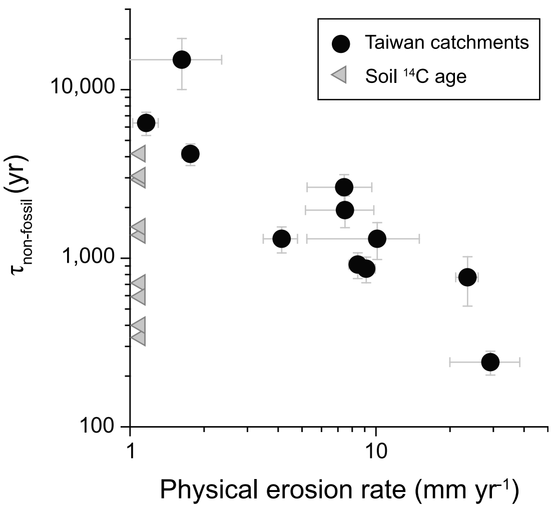

Figure 6. The time available for $POC_{non\text{-}fossil}$ aging imposed by physical erosion ($\tau_{non\text{-}fossil}$, yr) as a function of physical erosion rate (mm yr^{-1}) calculated from suspended sediment yields for catchments in Taiwan (circles). Triangles indicate the measured ^{14}C-age of surface soils (A-E horizons) in the Central Range [*Hilton et al.*, 2008a].

stock in vegetation and soil. Across Taiwan, the maximum $\tau_{non\text{-}fossil}$ imposed by physical erosion is ~15,000 yr in the Hsiukuluan River (Figure 6). Given the dominant role of respiration to carbon loss in terrestrial ecosystems, the estimates of $\tau_{non\text{-}fossil}$ are not directly comparable to estimates of residence time in vegetation and soil. These account for all input and output fluxes and recognize different pools of carbon which turnover at different rates [*Trumbore*, 1993]. However, the limit on biomass aging set by $POC_{non\text{-}fossil}$ export is consistent with the range of conventional radiocarbon ages of surface soils (A-E Horizons) in Taiwan, which reach a maximum of 4169 yr [*Hilton et al.*, 2008a] with the majority falling between 340 and 1540 yr (Figure 6).

[30] The data from Taiwan suggest that suspended sediment yields of ~3000–4000 t km^{-2} yr^{-1} (physical erosion rates of ~1–2 mm yr^{-1} with sediment density of 2.5 t m^{-3}) can limit $\tau_{non\text{-}fossil}$ to ~8000 yr (Figure 6). Thus, it appears that even modest rates of physical erosion can reduce or even eliminate the potential for very long timescales (>10,000 yr) available for pools of organic matter in soils to age, regardless of their respiration rate [*Trumbore*, 1993; *Torn et al.*, 1997]. $POC_{non\text{-}fossil}$ export thus plays an important role in montane ecosystem turnover, likely to promote young sections of forest where net productivity is most efficient [*Restrepo et al.*, 2009] and inhibit ecosystem retrogression [*Wardle et al.*, 2004; *Peltzer et al.*, 2010]. Physical erosion rates of 1–2 mm yr^{-1} are exceeded in many mountain belts [*Galy and France-Lanord*, 2001; *Dadson et al.*, 2003; *Hicks et al.*, 2004b; *Gabet et al.*, 2008; *Milliman and Farnsworth*, 2011] suggesting that erosion may limit $\tau_{non\text{-}fossil}$ in mountain forest at the global scale. At very high erosion rates of >10 mm yr^{-1}, the physical processes impose a timescale for aging (Figure 6) which encroaches on the centennial rates of turnover in vegetation and components of soil organic carbon [*Trumbore*, 1993; *Torn et al.*, 1997]. Clearly, the findings here demonstrate that the impact of rapid geomorphic process rates on nutrient and carbon cycling in mountain forests warrants further assessment.

5.4. Wider Implications for the Carbon Cycle

[31] The erosion and export of $POC_{non\text{-}fossil}$ by mountain rivers represents a lateral flux of recently fixed atmospheric CO_2 and its fate is important for our understanding of the global carbon cycle [*Berner*, 1982; *Hayes et al.*, 1999]. If this material is buried in sedimentary deposits while the $POC_{non\text{-}fossil}$ is replaced by new primary productivity on land, then this transfer represents a net sink of atmospheric CO_2. Efficient burial of $POC_{non\text{-}fossil}$ offshore Taiwan may be driven by the very high suspended sediment loads of the mountain rivers which deliver ~380 × 10^6 t yr^{-1} to the ocean [*Dadson et al.*, 2003], causing rapid accumulation rates in depocenters, a first order control on organic carbon burial efficiency [*Canfield*, 1994; *Galy et al.*, 2007]. Hyperpycnal river plumes, arising when SSC>40 g L^{-1} at the river mouth [*Mulder and Syvitski*, 1995], can trigger turbidity currents which are also thought to play an important role by rapidly delivering $POC_{non\text{-}fossil}$ carried by floodwaters (Figure 1b) to deep marine sediments [*Dadson et al.*, 2005; *Kao et al.*, 2006; *Nakajima*, 2006; *Saller et al.*, 2006; *Hilton et al.*, 2008a]. While the fate of $POC_{non\text{-}fossil}$ remains to be fully assessed, it seems likely that a large proportion of the 0.5 ± 0.2 × 10^6 tC yr^{-1} of $POC_{non\text{-}fossil}$ delivered to the oceans from Taiwan is buried.

[32] The significance of the transfer of $POC_{non\text{-}fossil}$ from Taiwan to the ocean is evident from comparison to a well-studied source-to-sink region from the Himalayan mountain belt to Bay of Bengal. There, an estimated 3.7 × 10^6 tC yr^{-1} of $POC_{non\text{-}fossil}$ is delivered by the Ganga-Brahmaputra rivers and sequestered from a continental source region ~50 times larger than Taiwan [*Galy et al.*, 2007]. The conservative estimate of $POC_{non\text{-}fossil}$ flux from the small mountain island represents ~15% of this value and ~1% of the estimated total terrestrial organic carbon burial in the oceans [*Schlünz and Schneider*, 2000]. Evidently, mountain islands are important not only for the erosion and transfer of POC_{fossil} [*Blair et al.*, 2003; *Leithold et al.*, 2006; *Kao et al.*, 2008; *Hilton et al.*, 2011a], but also in the transfer of carbon recently fixed from atmospheric CO_2.

[33] Our data suggest that, for a constant set of geomorphic conditions, the fluvial transfer of $POC_{non\text{-}fossil}$ from mountain catchments is driven by climate (Figure 3) through the activation of erosion and transport processes during heavy rainfall (Figure 1b). A move to a wetter, stormier climate over mountain forest should enhance the erosional export of $POC_{non\text{-}fossil}$. In settings with strong coupling between depositional sinks and terrestrial inputs [e.g., *Leithold and Hope*, 1999; *Kao et al.*, 2006] this offers a feedback in the Earth System, whereby climate modifies rates of carbon sequestration through erosion and burial of $POC_{non\text{-}fossil}$ [e.g., *Hilton et al.*, 2008a]. In addition, the data from Taiwan suggest that this carbon transfer is moderated by the catchment geomorphology (Figures 4a and 5). Rapid rates of plate convergence and the uplift of competent metamorphic rocks set prime conditions for the rapid erosion and fluvial export of $POC_{non\text{-}fossil}$ concomitant with large amounts of clastic sediment [*Galy et al.*, 2007; *Hilton et al.*, 2008a, 2008b]. On orogenic timescales, this implies a tectonic forcing of the carbon cycle which may lead to net

changes in the size of the organic carbon reservoir and influence atmospheric greenhouse-gas concentrations [*Derry and France-Lanord*, 1996; *France-Lanord and Derry*, 1997; *Hayes et al.*, 1999] via a carbon transfer that is sensitive to climatic conditions [cf. *West et al.*, 2005].

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